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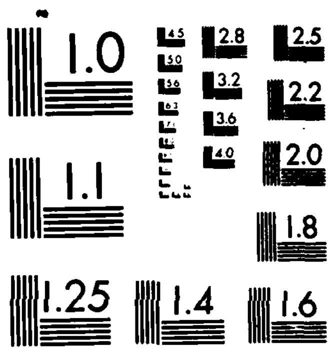

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OPERATIONAL SOIL MOISTURE PREDICTION MODEL

Final Technical Report

by

Lev N. Gutman

December 1985

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ABSTRACT

The one dimensional problem of the diurnal variations of moisture and temperature at different depths of the thermally active, non-frosen, non-swelling soil layer is considered.

It is assumied that we know: (1) values of all meteorological elements at anemometer level at any time, (2) Moisture and temperature distribution with depths at some initial moment, and (3) Physical parameters and caracteristics of the soil and vegetation cover which we shall call canopy for brevity.

The model consists of: (1) Partial differential equations of the subsurface hydrology for a non-saturated vapor-liquid flow in the soil, allowing for the moisture removal by the roots of plants. (2) Algebraic empirical and ordinary differential equations ~~suggested by Deardorff (1978)~~ for the description of the radiative and turbulent heat transfer between the atmosphere and the soil through the canopy allowing for the influence of condensation or evaporation at the foliage and also the retention of a part of the precipitation by its the foliage. (3) Moisture and vapor balance equations at the surface of the soil. (4) Continuity conditions for heat and vapor fluxes at the top of the canopy. (5) Semiempirical algebraic equations of the type of surface layer equations for a thin atmospheric layer between the canopy and anemometer level. to p. 4

The method of solution is the following:

1. Transforming and solving the equation describing the effect of canopy, we construct boundary condition for moisture and temperature at the soil surface which takes into account that effect. 2. The soil temperature and moisture equations as well as the boundary and initial conditions are approximated by the finite-difference equations in accordance with known Crank-Nicolson schemce

respectively. These schemes are quite simple and provide the convergence and the second order of the accuracy with respect to the time and space steps. 3. The finite-difference equations are solved by the tridiagonal algorithm in combination with iteration method. fr. p. i

The solution has been programmed. Resultes of computations are in agreement with experimental data. A few examples of soil moisture computations are presented together with experimental data.

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1. INTRODUCTION

Information on soil moisture and temperature at the surface and at different depths is of interest in numerous fields of human activity, e.g., underground structures (water pipes, sewage, electric and other cables), road building, some military problems (for example, road practicability for tanks), etc. In agriculture and agrometeorology, this information is often very necessity. For semiarid and arid zones such information becomes especially important.

There is usually a great deal of technical difficulty and expense involved in conducting special soil moisture and temperature measurements. Therefore, it is very tempting to have a model which would allow the computation of soil moisture and temperature from routine meteorological observations available for many regions of the globe. Moreover, such a model would enable the prediction of soil moisture and temperature using the meteorological forecast.

The construction of a model encounters some difficulties such as inhomogeneities of the land surface and of the soil at various depths, lack of information on physical parameters and presence of different urban constructions and artificial covers (such as concrete), as well as natural vegetation cover.

It has been shown that vegetation can significantly influence the heat and moisture transfer between the earth's surface and the atmosphere. Therefore, since most of the land surface of the globe is covered by vegetation, it is desirable to include the vegetation effect in a model which simulates the heat and moisture transfer at the land surface.

The object of the present paper is to construct a model for an operational soil moisture and temperature computation from meteorological data, taking into

account the effect of the vegetation cover which hereafter will be referred to as a canopy. To avoid the other difficulties mentioned above, we assume that the soil is homogeneous, and the land surface is horizontal and free of urban constructions.

To our knowledge, there are no publications devoted even to such a simplified problem, although there are many papers in which soil moisture and/or soil temperature computations have been carried out. These works have been directed at the investigation of soil physics or of a methodology of numerical solution (e.g., [1], [2]). In these papers, the boundary conditions at the surface are mostly simplified, e.g., evaporation from the surface is assumed constant and known. Such assumptions are not sufficient for our problem. Included in this group are some papers in which an attempt is made of coupling the soil part of the problem with a rather primitive atmospheric model assuming bare soil with attention directed mainly at the soil (e.g., [3], [4]). These works have allowed us to arrive at the important conclusion that, in principle, soil moisture and temperature computation, with the accuracy necessary for many aims, is possible.

Another group of papers is devoted to meteorological and climatological problems in which the soil moisture and temperature play an auxiliary role. In the majority of these works the soil part of the problem is highly simplified (e.g., [5]) or, if not oversimplified, cumbersome calculations and the lack of necessary information make it very difficult to reliably estimate the accuracy with which soil moisture is determined (e.g., [6]). In this group of papers, we refer only to those in which attempts to allow for the vegetational cover effect have been made. Obviously, canopy models accessible for inclusion into complex models containing soil and atmospheric parts, must be sufficiently simple with

respect to mathematics. Nevertheless, there have been attempts to construct rather complicated two-level, or even multi-level, biosphere models for inclusion into general circulation models of the atmosphere [7]. Deardorff (1978) [5] has suggested a single-level canopy model, sufficiently simple with respect to mathematics and at the same time rather comprehensive with respect to physics. The model is based on moisture and heat balance equations for the vegetation layer and for the surface of the earth and on some empirical relationships. Deardorff's model has already been employed by some authors ([6], [8]). In the present paper we also use Deardorff s model while introducing minor modifications for its improvement.

Unlike Deardorff and the authors mentioned above, we have made some preliminary analytical transformations in an attempt to simplify the solution of the vegetational part of the problem. The present paper includes only the physical and mathematical formulation of the problem and its method of solution. Results of computations and comparisons with observations, as well as physical conclusions, will be published separately.

2. NOTATION

Arabic

a	anemometer level height
b_T	empirical value dependent on type (textural class) of soil only
C	volumetric heat capacity of soil
C_f	heat or moisture transfer coefficient for the foliage element
C_{Ho}	heat and moisture transfer coefficient applicable to bare soil. C_{Hg}: to soil under a canopy, C_{Hh}: to the top of a dense canopy.
D_θ	moisture diffusivity of soil
D_T	thermal moisture diffusivity of soil
E	evaporation rate, if $E>0$, or condensation rate is $E<0$
E_{tr}	transpiration rate (per unit soil surface area)
g	gravity acceleration
h	thickness of the vegetation layer
H	sensible heat flux, positive upwards
i,j	depth and time indices
K	hydraulic conductivity of soil
L	latent heat of vaporization
L_r	length of roots per unit soil volume
N	number of levels in soil
P	precipitation rate
Q	soil heat flux density, positive downward
q	specific humidity
q_s	saturated humidity
r	fraction of potential evaporation rate from foliage

r_a atmospheric resistance

r_c resistance coefficient dependent upon plant type

r_s generalized stomatal resistance

R surface runoff

R_L downward directed longwave radiative flux

R_S downward directed shortwave radiative flux

R_{Smax} maximum value of R_S

R_w gas constant for water vapor

R_{iB} bulk Richardson number

t time

T temperature

u wind speed

W flux density of soil water transfer, positive downwards

z vertical coordinate positive downwards (depth in soil)

z_ℓ lower boundary of the solution domain

z_r depth of root zone

Greek

α_f foliage albedo

α_g ground surface albedo

$\Delta z, \Delta t$ depth and time increments

ϑ volumetric moisture content

ϑ_d residual water content

ϑ_{dew} mass of liquid water retained by foliage per unit horizontal area

ϑ_{dmax} maximum value of ϑ_{dew}, beyond which runoff to soil occurs

ϑ_{rmin} minimum value of ϑ in the root zone

ϑ_s value of ϑ at saturation

ϑ_{wilt} wilting point value of

ρ density of water

ρ_a density of air

λ thermal conductivity of soil

ψ soil moisture potential

σ Stefan-Boltzman constant

σ_c cloud fraction

σ_f foliage shielding factor of ground from shortwave radiation (area average)

ε_f foliage emissivity

ε_g ground surface emissivity

Subscripts

a reference "anemometer level" height

f foliage surface

g value at the ground surface

af mean value within a canopy

h value at the top of the canopy

s value for saturated soil

3. FORMULATION OF THE PROBLEM

We consider the problem of diurnal moisture and temperature variations in the upper thermally active soil layer which we assume to be non-swelling. In addition, we assume that we know the values of the wind, temperature, pressure, humidity, precipitation, direct and scattered radiation at the shelter level at all times and that we have all the necessary information about the physical parameters of the soil and the canopy. We imply that there is a layer of air between the shelter and the canopy.

To describe the heat and moisture transfer in the soil, we employ a system of subsurface hydrology equations for unsaturated vapor-liquid flow [9], [10]. We assume that nothing depends on the horizontal coordinates and that the ground surface is horizontal. The soil moisture equations will be supplemented by the water extraction term [11] which parameterizes the moisture removal from the soil by the roots of the plants.

As a result, the soil-moisture equations are reduced to the following one-dimensional, non-stationary system of partial differential equations.

$$\frac{\partial \vartheta}{\partial t} = -\frac{\partial}{\partial z}\left(W/\rho\right) - S \tag{1}$$

$$C\frac{\partial T}{\partial t} = -\frac{\partial Q}{\partial z} \tag{2}$$

where

$$W/\rho = -D_\theta \frac{\partial \vartheta}{\partial z} + K - D_T \frac{\partial T}{\partial z} \tag{3}$$

$$Q = -\lambda \frac{\partial T}{\partial z} \tag{4}$$

In these equations the z axis and fluxes are positive downward.

Unlike [9], eq. (4) does not take into account the transfer of latent heat by vapor movement induced by the moisture gradient. The possibility of omitting this effect can be shown by elementary estimation. The fact is that the thermal vapor diffusivity coefficient becomes extremely small for soil moisture values observed in nature (see e.g., Fig. 2 from [9]).

Neglecting the effect of hysteresis, one can consider parameters C, D_ϑ, D_T, K, λ and also ϑ as unique functions of ψ and of the textural class (or type) of the soil, whereas parameters ρ, ϑ_s, ϑ_d depend only on textural class of the soil. Following [12]-[15] we set*)

$$C = C_s(1-\vartheta_s-\vartheta), \quad \psi = \psi_s(\vartheta_s/\vartheta)^{b_T}, \quad K = K_s(\vartheta/\vartheta_s)^{2b_T+3}$$

$$D_\vartheta = (-\psi_s) b_T K_s \vartheta_s^{-1} (\vartheta/\vartheta_s)^{b_T+2} \tag{5}$$

$$\lambda = 2.57 \cdot 10^{-4} + 0.062(-\psi)^{-1.04} \; (cal/sec \cdot cm \cdot {}^\circ C)$$

The values ϑ_s, C_s, ψ_s, K_s and b_T for different soil are given in Table 2 from [12].

To parameterize the soil moisture removal by roots, we adopted an expression for S, suggested recently by [11] (see also [16]).

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*) The expression for λ approximates quite accurately the experimental data plotted in Fig. 4 from [13].

$$S = \mathcal{L}_r(z) K E_{tr} \Big/ \int_0^{z_r} \mathcal{L}_r(z) K dz \qquad (6)$$

assuming that $\mathcal{L}_r(z)$ and z_r are known.

We now proceed to the formulation of boundary and initial conditions of the problem.

Since the ground surface is assumed to be horizontal, one can, with slight simplification, consider that surface runoff occurs only if the precipitation water has no time to infiltrate into the soil and to evaporate from the ground surface, which becomes saturated instantly:

$$\left.\begin{array}{lll} \text{either } \vartheta = \vartheta_s & \text{and } R > 0 \\ \text{or } \quad \vartheta \leq \vartheta_s & \text{and } R = 0 \end{array}\right\} \text{ at } z = 0 \qquad (7)$$

where

$$R = P_g - E_g - W_g \qquad (8)$$

$$P_g = \begin{cases} (1-\sigma_f)P & \text{if } \quad \vartheta_{dew} < \vartheta_{dmax} \\ P & \text{if } \quad \vartheta_{dew} = \vartheta_{dmax} \end{cases} \qquad (9)$$

(7) is actually an upper boundary condition for the soil moisture. (8) is the surface moisture balance equation. (9) allows for the retention of part of the precipitation by the foliage. As an upper boundary condition for the soil temperature, we shall use the ground surface heat balance equation

$$Q=(1-\sigma_f)\left[(1-\alpha_g)R_s+\varepsilon_g R_L\right]-\varepsilon_2 T_g^4+\sigma_f\varepsilon_1 T_f^4-H_g-LE_g \tag{10}$$

where

$$\left.\begin{aligned}
&R_L=\left[\sigma_c+1.21(1-\sigma_c)q_a^{0.08}\right]\sigma T_a^4\\
&\alpha_g=\begin{cases}0.31-0.34\,\vartheta_g/\vartheta_s & \vartheta_g<0.5\vartheta_s\\ 0.14 & \vartheta_g\geq 0.5\vartheta_s\end{cases}\\
&\varepsilon_1=\varepsilon_f\varepsilon_g\sigma/(\varepsilon_f+\varepsilon_g-\varepsilon_f\varepsilon_g)\\
&\varepsilon_2=(1-\sigma_f)\sigma\varepsilon_g+\varepsilon_1\sigma_f
\end{aligned}\right\} \tag{11}$$

We assume that R_s is known for any time from observations. It is also possible to use astronomic formulas for R_s, as in [6].

Eq. (10) contains additional terms, which take into account the effect of the canopy. If $\sigma_f = 0$, eq. (10) becomes the well-known ground surface heat balance equation for the bare soil.

Equations which have to parameterize the influence of the canopy and meteorological conditions on the heat and moisture transfer between the atmosphere and the ground are:

1. Canopy energy balance equation

$$\left.\begin{aligned}
&\sigma_f\left[(1-\alpha_f)R_s+\varepsilon_f R_L+\varepsilon_1 T_g^4-\varepsilon_3 T_f^4\right]=H_f+LE_f\\
&\qquad(\varepsilon_3=\varepsilon_1+\sigma\varepsilon_f)
\end{aligned}\right\} \tag{12}$$

where

$$\left.\begin{aligned}
H_g &= \rho_a c_p C_{H_g} u_{af} (T_g - T_{af}) \\
H_f &= \sigma_f \rho_a c_p C_u (T_f - T_{af}) \\
E_g &= \rho_a C_{H_g} u_{af} (q_g - q_{af}) \\
E_f &= \sigma_f \rho_a C_u (q_f - q_{af}) \\
C_{H_g} &= (1 - \sigma_f) C_{H_0} + \sigma_f C_{H_h}
\end{aligned}\right\} \tag{13}$$

$$\left.\begin{aligned}
u_{af} &= (1 - \sigma_f + 0.83 \sigma_f C_{H_h}^{1/2}) u_a \\
C_u &= 7.7 \cdot 10^{-2} (u_{af} + 0.3\ m/sec)
\end{aligned}\right\} \tag{14}$$

2. The relationship between the moisture characteristics at the foliage surface and the in-canopy air humidity.

$$q_f - q_{af} = r[q_s(T_f) - q_{af}] \tag{15}$$

where

$$r = 1 - \delta_c \frac{1 - \xi^{2/3}}{1 + \xi} \quad \left(\xi = \frac{\vartheta_{dew}}{\vartheta_{dmax}}\right) \tag{16}$$

$$\delta_c = \begin{cases} 0, \text{if there is condensation onto the foliage surface} \\ \quad q_{af} > q_s(T_f) \\ \\ 1, \text{if there is evaporation from the foliage surface} \\ \quad q_{af} \leq q_s(T_f) \end{cases} \tag{17}$$

$$\xi = \frac{r_a}{r_s} = \frac{1}{r_c C_f u_{af}} \left[\frac{R_{smax}}{R_s + 0.3 R_{smax}} + \left(\frac{\vartheta_{wilt}}{\vartheta_{rm}} \right)^2 \right]^{-1} \tag{18}$$

$$q_s(T) = 0.38 \cdot 10^{-2} \exp\left(17.269 \, \frac{T - 273.16}{T - 35.86}\right) \tag{19}$$

This empirical formula is actually one possible version of the well-known Clausius-Clapeyron equation.

3. The conservation equation for ϑ_{dew}:

$$\frac{d\zeta}{dt} = \frac{1}{\vartheta_{dmax}} \left[\sigma_f P - (E_f - E_{tr}) \right] \quad (0 \leq \zeta \leq 1) \tag{20}$$

where

$$E_{tr} = \delta_c \cdot \xi \cdot (1 - \zeta^{2/3}) \cdot E_f \Big/ (1 + \xi) \cdot r \tag{21}$$

4. Continuity conditions of the heat and water vapor fluxes at the top of the canopy layer

$$H_h = H_g + H_f \qquad E_h = E_g + E_f \tag{22}$$

where

$$\left.\begin{aligned} H_h &= \rho_a c_p C_{H_h} u_a (T_h - T_a) \\ E_h &= \rho_a C_{H_h} u_a (q_h - q_a) \end{aligned}\right\} \tag{23}$$

We denote

$$\left.\begin{aligned} T_h &= (1-\sigma_f) T_g + \sigma_f T_{af} \\ q_h &= (1-\sigma_f) q_g + \sigma_f q_{af} \end{aligned}\right\} \tag{24}$$

The last two relationships are constructed in accordance with an idea of the simplest linear interpolation with σ_f, also suggested in [5].

5. Continuity condition for q at the ground surface, g.

$$q_g = q_s(T_g) \exp(g \psi_g / R_w T_g) \tag{25}$$

6. Semiempirical relationship for C_{H_h} (see [8])

$$C_{H_h} = \begin{cases} 4.2 \cdot 10^{-3} \left[1 + 24.5 \left(-4.2 \cdot 10^{-3} R_{i_B}\right)^{1/2}\right], & \text{if } R_{i_B} < 0 \\ 4.2 \cdot 10^{-3} / (1 + 11.5 R_{i_B}), & \text{if } R_{i_B} \geq 0 \end{cases}$$

where

$$\left.\begin{aligned} R_{i_B} &= \frac{g(a - \sigma_f h)}{u_a^2 + u_c^2}\left(1 - \frac{T_h}{T_a}\right) \\ u_c &= \begin{matrix} 0.1\ m/sec & \text{if } T_h < T_a \\ 1.1\ m/sec & \text{if } T_h \geq T_a \end{matrix} \end{aligned}\right\} \tag{26}$$

These relationships imply that a thin constant flux layer exists between the canopy and the anemometer level.

For the formulation of the lower boundary conditions for ϑ and T, we assume that at a certain depth (which is usually 1-2 m), diurnal variations of ϑ and T can be neglected. We consider two cases: the soil below that depth is dry; the soil below that depth is saturated (water table):

Thus, the lower boundary conditions for ϑ and T are

$$\left.\begin{array}{l} \vartheta = \vartheta_{\ell} = \text{const (dry soil)} \\ \text{or} \\ \vartheta = \vartheta_{s} = \text{const (saturated soil)} \\ \\ T = T_{\ell} = \text{const} \end{array}\right\} \text{ at } z = z_{\ell} \tag{27}$$

The values ϑ_{ℓ}, ϑ_{s}, T_{ℓ} and z_{ℓ} are assumed to be known from climatic and edaphic data.

To close the problem, one needs to formulate the initial conditions.

We assume that, at a given moment of time $t = 0$, ϑ and T are given functions of z:

$$\vartheta = \vartheta_{o}(z) \ , \quad T = T_{o}(z) \quad (o < z < z_{\ell}) \tag{28}$$

Eq. (20) also requires an initial condition for ζ. Evidently it is difficult to obtain information concerning liquid water on the foliage. Therefore we set

$$\zeta = 0 \quad \text{at} \quad t = 0 \tag{29}$$

assuming that our computation starts at the moment when all the water retained by the foliage after the last rain has evaporated.

4. SOLUTION OF THE PROBLEM

First of all we introduce a finite-difference time grid $t_j = j\Delta t$, $\varphi_j = \varphi(z, t_j)$ $j = 1,2,\ldots$, where φ is any function or parameter of the problem (including those which are independent of z).

To clarify the explanation of the computation scheme, we shall assume that φ_{j-1} are known. We then describe the iterations involved in advancing to the next time step. This procedure will then be used to continue for all j . For brevity we shall omit the index when it is j . This notation permits us to regard eqs. (12)-(19), (21)-(24), as written for the moment t_j .

Eq. (20) can be approximated by the following finite-difference form

$$\frac{\zeta - \zeta_{j-1}}{\Delta t} = \frac{1}{\vartheta_{dmax}} \left(\sigma_f P + E_{tr} - E_f \right) \quad (0 \leq \zeta \leq 1) \tag{30}$$

This implicit scheme is chosen in order to refer dependent variables in (30) to the moment t_j , as in eqs. (12)-(19), (21)-(24).

Considering the crudeness of eq. (20), the first order approximation of the time derivative in (30) appears to be sufficient.

Let us assume temporarily that T_g and ϑ_g (i.e., $(T_g)_j$ and $(\vartheta_g)_j$) are known. Then, as one can show, (12)-(19), (21)-(24) together with (30) is a closed algebraic system from which we shall find T_f in the following way.

First, substituting (13), (23) and (24) into (22) and making some simple transformations, we obtain

$$q_{af} = a_a q_a + a_f q_f + a_g q_g \tag{31}$$

$$T_{af} = a_a T_a + a_f T_f + a_g T_g \tag{32}$$

where

$$\left.\begin{aligned} a_a &= c_{H_h} u_a \Big/ \left[c_{H_g} u_{af} + \sigma_f (c_u + c_{H_h} u_a) \right] \\ a_f &= \sigma_f c_u \Big/ \left[c_{H_g} u_{af} + \sigma_f (c_u + c_{H_h} u_a) \right] \\ a_g &= 1 - a_a - a_f \end{aligned} \right\} \tag{33}$$

In the right hand sides of (31) and (32) all quantities with the exception of q_f and T_f, are known. Indeed q_a and T_a are known from observations, T_g and q_g are known, as a consequence of our assumption and eq. (25) respectively.

Concerning a_a, a_f and a_g, we note the following. In [5] it was assumed that c_{H_h} is known constant. We calculate c_{H_h} from (26), but in ( ) we shall use $(T_h)_{j-1}$. Then a_a, a_f and a_g can be calculated with the aid of (33). It should be noted that (31) and (32) are similar to relationships suggested in [5], where a_a , a_f and a_g were assumed to be given functions of σ_f alone. Eqs. (15) and (31) can be solved for q_f and q_{af} . Substituting q_f, q_{af}, obtained in this way, into the expression for E_f from (13), we get

$$E_f = \left[q_s(T_f) - q_{ag} \right] r \sigma_f a_E \Big/ (1 - a_f + r a_f) \tag{34}$$

where

$$q_{ag} = (a_a q_a + a_g q_g) \big/ (1 - a_f) \ , \quad a_E = \rho_a c_u (1 - a_f) \tag{35}$$

Note, that $0 \leq r \leq 1$ and $0 \leq a_f \leq 1$. Therefore the sign of E_f depends only on

the sign of the difference in the brackets in (34).

Substituting (34) into (12) yields an equation, which links T_f with r:

$$F(T_f, r) = \frac{r}{1 + r a_f - a_f}\left[q_s(T_f) - q_{ag}\right] - b(T_f) = 0 \tag{36}$$

where

$$b(T_f) = \frac{c_p}{L}(T_{ag} - T_f) + \frac{1}{L a_E}(R_E - \varepsilon_3 T_f^4) \tag{37}$$

in which

$$R_E = (1 - \alpha_f) R_s + \varepsilon_f R_L + \varepsilon_1 T_g^4 \tag{38}$$

$$T_{ag} = (a_a T_a + a_g T_g)/(1 - a_f) \tag{39}$$

The expressions $F(T_f, r)$ and $b(T_f)$ can be treated as given functions of their arguments.

From (34) and (36) we obtain

$$E_f = \sigma_f a_E b(T_f) \tag{40}$$

In this formula, unlike (34), the explicit dependence E_f on r is absent.

Combining (13), (31), (35) and (40), we derive:

$$q_f = q_{ag} + b(T_f), \quad q_{af} = q_{ag} + a_f b(T_f) \tag{41}$$

Recalling the expressions for E_f: (13), (34) and (40), one can understand from (41) why the differences $q_s(T_f) - q$, $q_s(T_f) - q_{af}$ and $q_s(T_f) - q_{ag}$ always have the same signs.

Thus, from (34) and (36) it follows that the sign of $b(T_f)$ determines

whether condensation or evaporation is occuring at the foliage:

$$\left.\begin{array}{l} \text{If } \mathscr{E}(T_f) < 0 - \text{condensation} \\ \\ \text{If } \mathscr{E}(T_f) > 0 - \text{evaporation} \end{array}\right\} \tag{42}$$

On account of (16) and (17) in the case of condensation, T_f can be found from eq. (36) in which r = 1.

According to (17), and (21) E_{tr} = 0 in this case. Therefore, in the case of condensation (30) is reduced to

$$\zeta = \xi_{j-1} + (\sigma_f P - E_f)\Delta t / \vartheta_{dmax} \tag{43}$$

In the case of evaporation from the foliage, an equation for T_f can be obtained by excluding r and ζ from (16), (30) and (36).

Indeed, in this case, solving (16) and (36) for ζ and r respectively, and allowing for (41), we express ζ and r in terms of T_f:

$$\zeta(T_f) = [(\xi + 1) r(T_f) - \xi]^{3/2} \tag{44}$$

$$r(T_f) = (1 - a_f)\mathscr{E}(T_f) / [q_s(T_f) - q_{af}] \tag{45}$$

Substituting (40), (44) and (45) into (21) yields an expression for the transpiration rate

$$E_{tr} = \sigma_f a_E \xi [q_s(T_f) - q_f] / (1 - a_f) \tag{46}$$

which accompanies evaporation from the foliage.

If we substitute (40), (44) and (46) into (30), we arrive at an algebraic transcendental equation for T_f in the evaporation case

$$\zeta(T_f) - A \cdot F\left(T_f, \frac{\xi}{1+\xi}\right) - B = 0 \tag{47}$$

where the functions ζ and F are specified by (44) and (36) respectively and

$$\left.\begin{aligned} A &= \Delta t \cdot \rho_a c_u \sigma_f (1 - a_f + \xi) / \vartheta_{dmax} \\ B &= \zeta_{j-1} + \Delta t \cdot \sigma_f P / \vartheta_{dmax} \end{aligned}\right\} \tag{48}$$

Eq. (36) at $r = 1$, as well as at $r = \xi/(1+\xi)$, are solved numerically by the Newton-Raphson method [17]. Eq. (47) is solved using the so-called "rule of false position" [17].

As a result of (42), (44), (45) and (47) one can find T_f and determine whether condensation or evaporation at the foliage is occurring at the moment t_j.

Details are given in APPENDIX A.

Once we obtain T_f, then, using (13), (32), (37), (40), (41), (43)-(46), we can calculate the quantities:

$$\zeta, r, q_f, q_{af}, T_{af}, E_f \text{ and } E_{tr}, \tag{49}$$

which are necessary for the computation of the soil temperature T(z,t) and moisture ϑ(z,t).

Eliminating $W\rho$ and Q from(1)-(4) yields

$$C\frac{\partial T}{\partial t}=\frac{\partial}{\partial z}\lambda\frac{\partial T}{\partial z} \tag{50}$$

$$\frac{\partial \vartheta}{\partial t}=\frac{\partial}{\partial z}D_{\theta}\frac{\partial \vartheta}{\partial z}-\frac{\partial K}{\partial z}+\frac{\partial}{\partial z}D_{T}\frac{\partial T}{\partial z}-S \tag{51}$$

Note that these equations are written in their "prior to introducing the time indices" form.

Let us consider the soil temperature eq. (50). Sophocleous (1979) [18] solved the temperature equation in a more general form (including the moisture influence term). His estimates indicated that the dependence of the soil temperature on the soil moisture field is weak.

This fact justified the use of the simplified eq. (4) and also shows that changes of the parameters C and λ with ϑ only slightly affect T. This is reasonable because the changes of C and λ have to be mutually compensated to a certain extent.

The variables (49) depend on ϑ_g only through the soil moisture potential ψ_g, which appears in (25). It is easy to show that the exponential factor in (25) is close to unity, if the soil is even slightly wet.

However, in the case of extremely dry soil, the value of ϑ_g is very small. Therefore, in both cases the effect of ϑ_g on the values (49) is insignificant.

Thus, taking into account the weak dependence of the soil temperature field on the soil moisture, we shall use $(\vartheta_g)_{j-1}$ instead of $(\vartheta_g)_j$ for solving eq. (50).

To transform (10) into a convenient linear boundary condition for T, we

shall use the fact that the numerical procedure for solving eq. (50) contains a certain iteration process.

Let us turn to an approximate relationship

$$T_g^{\,4} = 4\left(T_g^{(n-1)}\right)^3 \cdot T_g - 3\left(T_g^{(n-1)}\right)^4 \tag{52}$$

which is valid when the temperature difference between two consecutive iterations is relatively small.

In (52) $T_g^{(n-1)}$ is taken from the (n - 1)th iteration. In analogy with the time indices, hereafter the iterative index in brackets will be omitted whenever it is (n) (i.e., referring to the n-th iteration). For example in (52) T_g means $(T_g)_j^{(n)}$.

Substituting (13) and (52) into (10), we arrive at a linear boundary condition for T at the ground surface.

$$-\lambda\frac{\partial T}{\partial z} + \nu T = \mu \qquad \text{at} \quad z = 0 \tag{53}$$

where

$$\left.\begin{aligned} \nu &= \rho_a c_p c_{H_g} u_{af} + 4\varepsilon_2\left(T_g^{(n-1)}\right)^3 \\ \mu &= (1-\sigma_f)\left[(1-\alpha_g)R_s + \varepsilon_g R_L\right] + 3\varepsilon_2\left(T_g^{(n-1)}\right)^4 + \\ &\quad + \sigma_f \varepsilon_1 T_f^{\,4} + \rho_a c_p c_{H_g} u_{af} T_{af} - L E_g \end{aligned}\right\} \tag{54}$$

Eq. (50) is rewritten in the finite differences in accordance with the wellknown Crank-Nicolson scheme. The boundary and initial conditions ((27),

(53) and (28)) are also presented in finite difference form. Such a finite-difference problem is solved at each iteration with the aid of the so-called tridiagonal algorithm (see e.g. [19]).

Details are presented in APPENDIX B.

Once $T(z,t)$ is known, we proceed to the solution of the soil moisture equation (51) with the boundary and initial conditions (7), (27) and (28).

For this we make use of the predictor-corrector equations (suggested by Douglas and Jones (1963) [20] and recommended in [19] for the type of problems under consideration) again solving them with the aid of tridiagonal algorithm at each iteration. By including a special logical scheme in the iteration process (see Fig. 4), we provide the satisfaction of the surface boundary conditions (7).

Details are presented in APPENDIX C.

Fig. 1 shows a simplified flowchart of the soil moisture and temperature calculation at given time step. In the following we explain it briefly. At the beginning all data, including those referred to the previous time step, are passed on from Box I to Box II. In Box II the equations for the vegetation layer are solved; in the first iteration, as a zero approximation, the values of the soil moisture and of the temperature at the surface are those calculated in the previous time step. Then the values of T_f, E_g, E_{tr} and H_g are passed on to Box III and / or to Box IV. In Box III the equation for the heat transfer is solved and a temperature profile is obtained. This is passed to Box IV. Successively in Box IV the equation for the soil moisture is solved and a moisture profile is obtained. Finally the values of the surface temperature and moisture calculated, respectivelly, in Box III and Box IV are passed on to Box II and a new iteration starts. At the end of the iterative process a new time

step in considered.

In Fig. 2 a scheme of the program which solves the vegetation layer equations is given. For the calculation of the transpiration rate and of the evaporation rate and the sensible heat flux at the ground surface some parameters have to be specified. They are listed in APPENDIX D where the listing of the program along with a table of the variables are presented.

In Fig. 3 a scheme of the program which solves the soil moisture flow equation without considering the effect of temperature gradients is given. For the calculation of the change with time of the soil moisture distribution an initial profile along with two boundary conditionsat each time step have to be specified. The boundary condition can be either the value of the soil moisture, pressure or the value of the moisture flux. In the latter case at the first step the value of the soil moisture presure at the boundaries has to be chosen in order to match the impposed flux. In APPENDIX E the listing of the program along with a table of the variables are presented.

For the tests some cases for wich there are experimental data and results of calculations carried out by means of some other methods have been picked out. The tests confirm that all Boxes work normally and that the soil moisture computation isaccomplished successfully. Some results of our soil moisture computations are presented in Fig. 5 and 7. They show the change with time of the soil moisture distribution with depth for two different boundary conditions and for the same values of the physical parameters.

Details on the physical parameters are given in the legends.

Some experimental data and numerical results obtained by (1) for the same case considered in Fig. 5, are presented in Fig. 6.

The comparison of Fig. 5 and Fig. 6 seems quite convincing. At last, Fig.8

illustrates how Box II works. It shows the result of our computation of the diurnal march of the foliage temperature. In the legend of Fig. 8 is given whole information about this run; some of the input values are taken from [5].

At present time we are continuing to work with the program. We are debugging the interaction between all four Boxes. Simultaneously we are preparing mass tests of the suggested method.

5. CONCLUSIONS AND RECOMENDATIONS

In conclusion we note the following. A method which allows to compute the distribution of the soil moisture and temperature with respect to the depth at any time is worked out. Also the moisture and heat fluxes from the soil, covered by the vegetation into the atmosphere can be computed.

The method needs the following information: data about the march of meteorological elements at the shelter level (wind, temperature, humidity, precipitation), data about the march of the cloudiness and solar radiation, data about some physical parameters of the soil and the vegetation cover whose thickness is assumed to be known and to be less than the height of the shelter level.

The method is based on the numerical solution of a system of equations describing the heat and moisture transfer in the soil and between the soil-vegetation system and the atmosphere. The numerical method which we use consists of the well known Crank-Nicolson predictor-corrector scheme in the form suggested by Douglas and Jones [19], [20], two internal and one external iteration processes. Such a method provides the necessary accuracy and convergency of the solution and it is rather economic from the point of view of the computer's time. At present time the programming is in act.*)

The theoretical part of the work was completed and accepted for publication in "Progress in Desert Research", proceedings of the Blaustein Institute for Desert Research (Israel) - UCLA (U.S.A.) Symposium, 1985.

*) It should be noticed that although the formulated problem is solved, however the work is not finished completely. It can be explained by the following. For some reasons which are beyond the author control he was almost deprived of a programmer or of any assistance during his work with the contract. Only in the last few months he had a programmer for one day work a week. At last, in December 1985 the author has received an assistence by a qualified scientst and programmer (M.G. Scarpino) who had an opportunity to make aquaintance with the problem during two months autumm 1985. At present time the work is being carried out quickly and it might be completed in a few months.

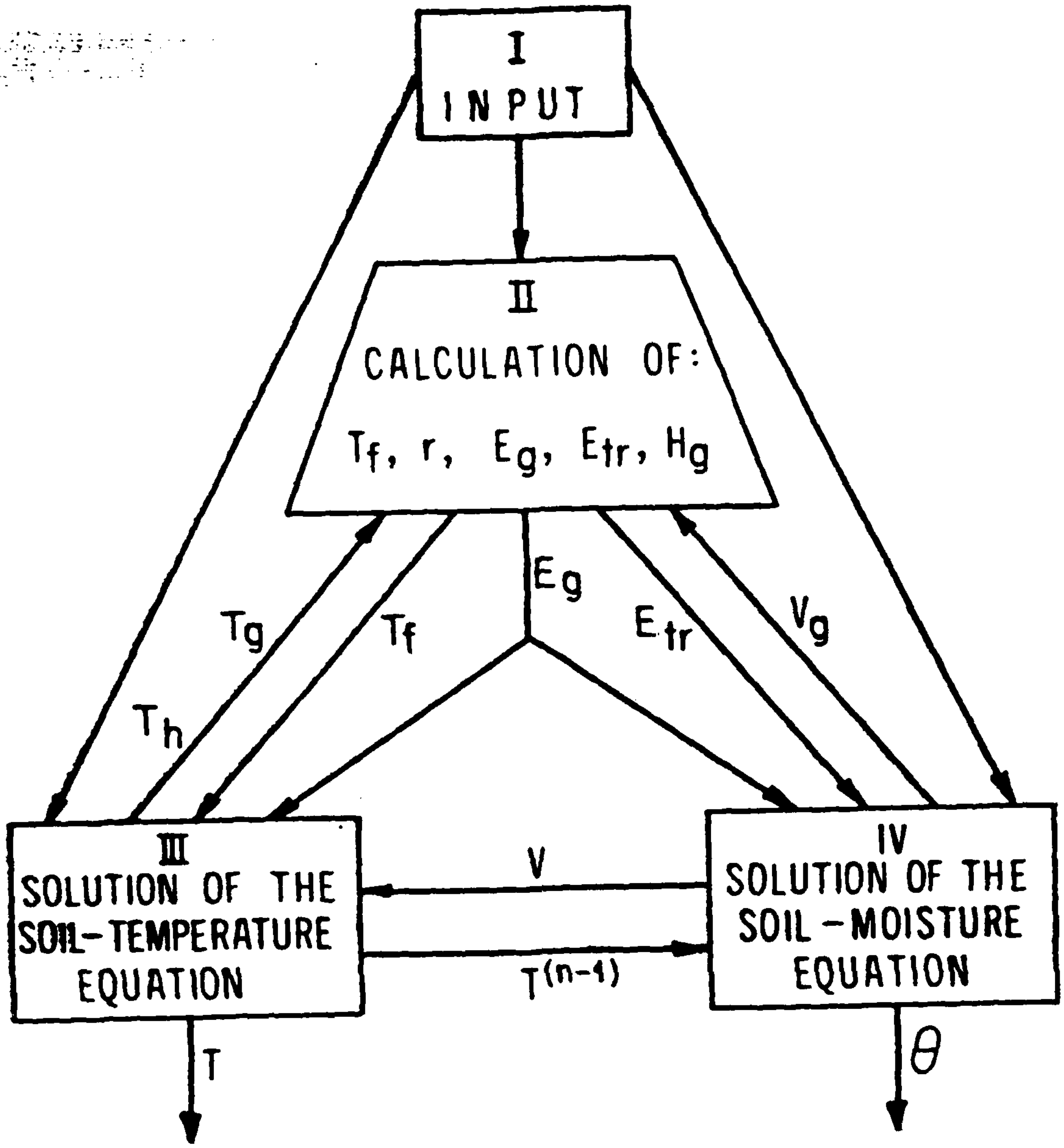

FIG. 1

Flowchart for soil moisture and temperature computation.

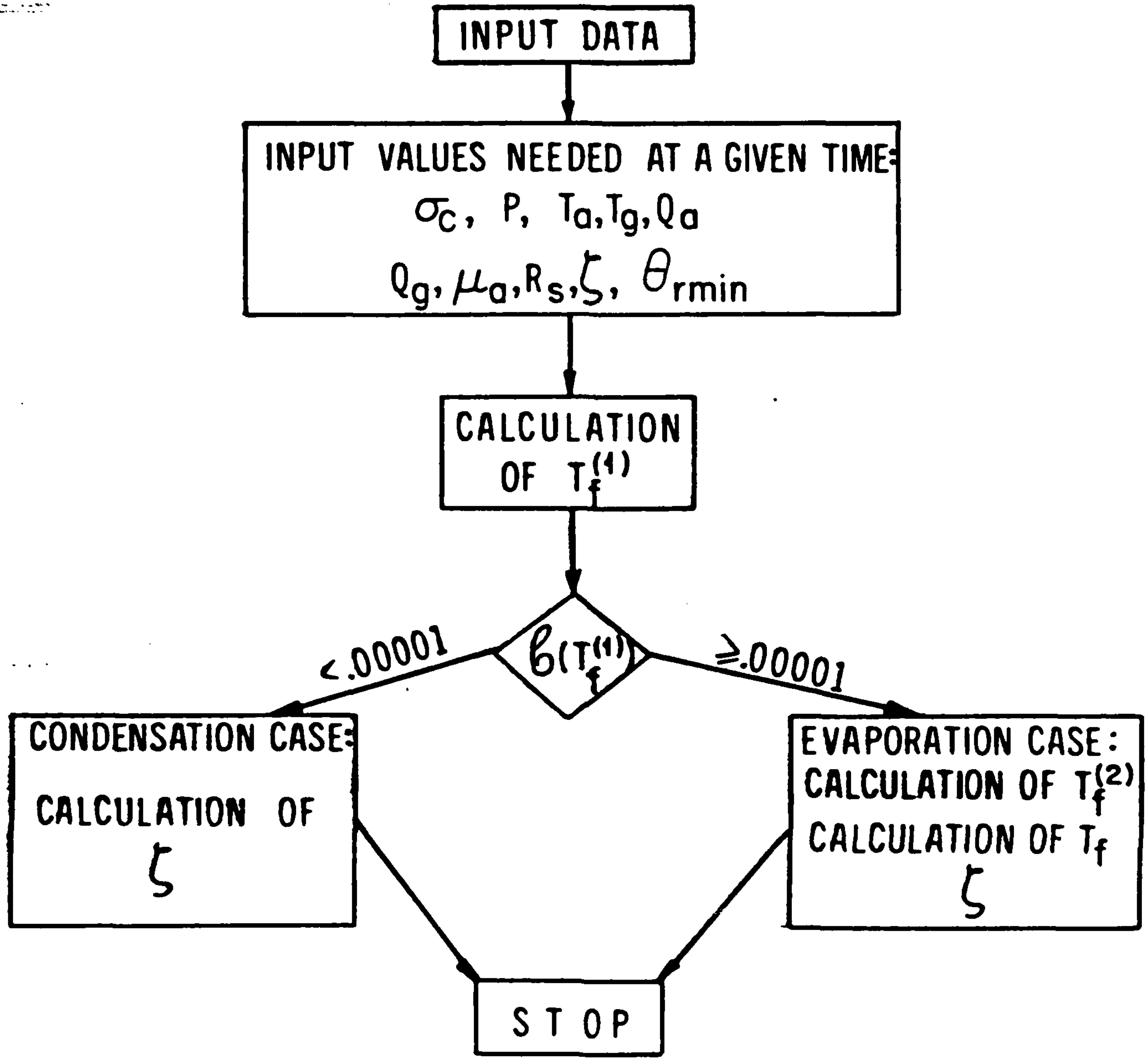

FIG. 2

Program Green scheme.

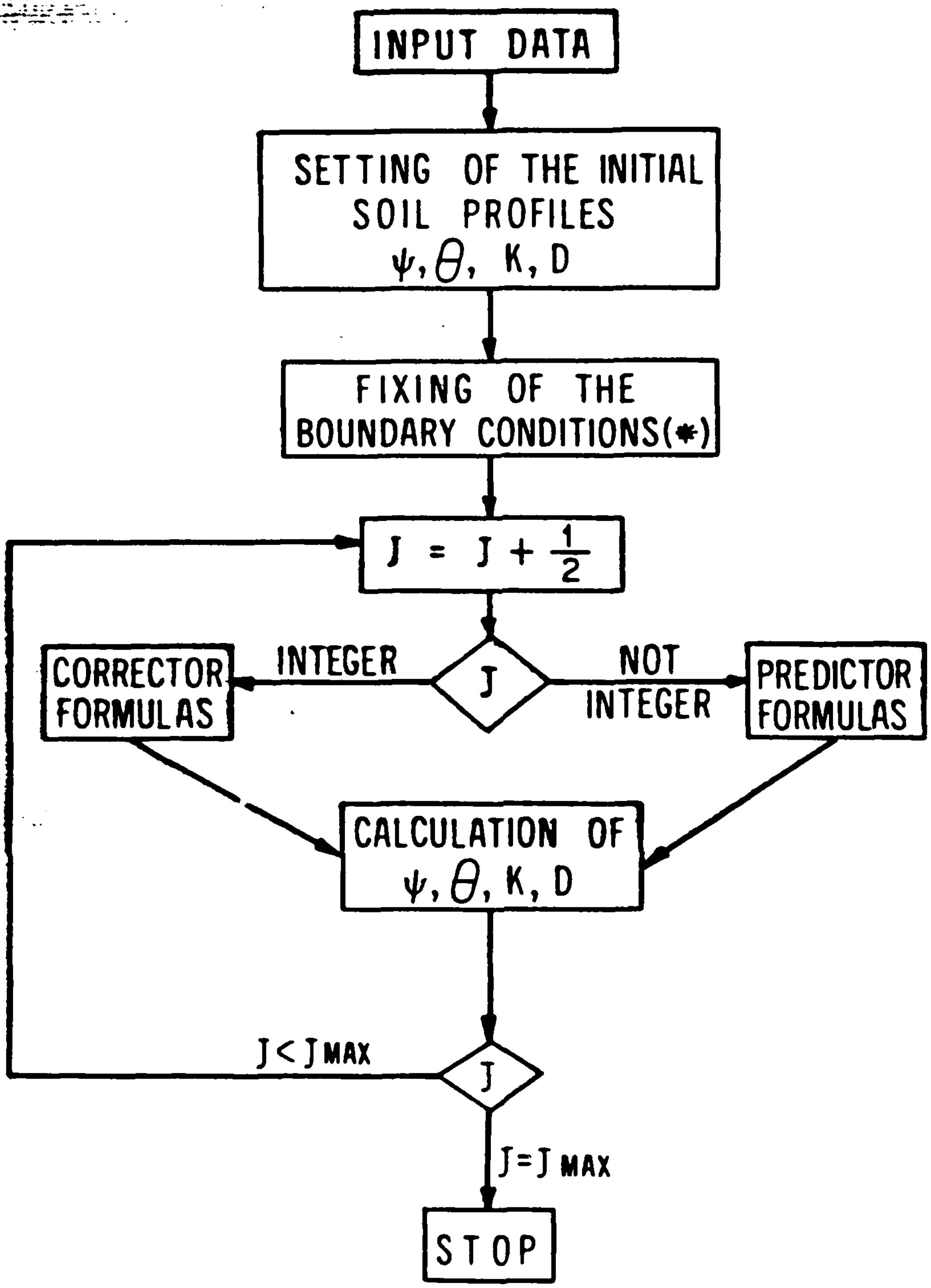

FIG. 3

Program Soil scheme. The program can run with two types of boundary conditions, namly with an imposed flux or with a fixed moisture content.

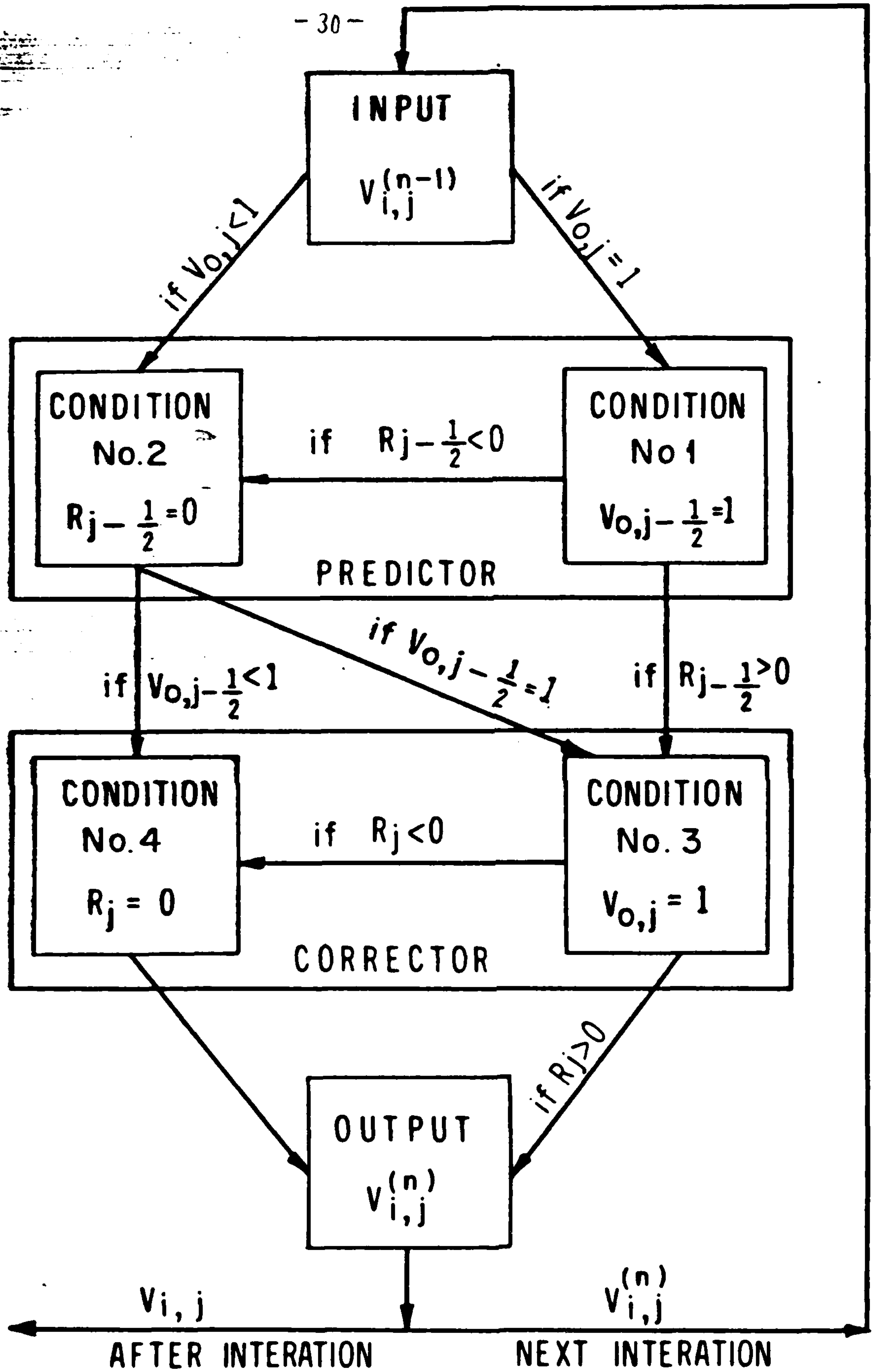

FIG. 4

A scheme of the calculation of ϑ satisfying to the boundary condition at the surface.

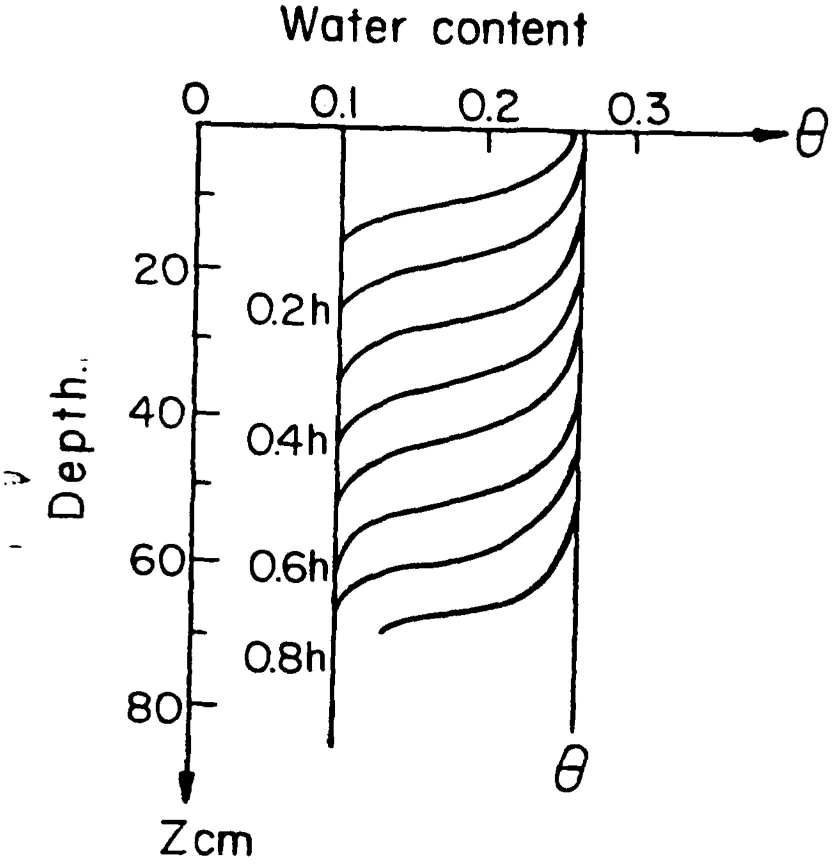

FIG.5

The distribution of ϑ with respect to the depth of the soil for t=0.1h, 0.2h, ..0.8h computed by our program for the problem: W/ρ =13.69cm/h at z=0, ϑ=0.1 at z=70cm, and at t=0. ($\Delta z = 1\,cm$, $t = 5\,sec$). It is assumed

$$K = K_s\left(1 + 0.85\cdot 10^{-6}|\psi|^{4.74}\right)$$

$$\vartheta = (\vartheta_s - \vartheta_d)\left(1 + 0.62\cdot 10^{-6}|\psi|^{3.96}\right) + \vartheta_d$$

where K = 34cm/h, ϑ_s=0.287, ϑ_d=0.075. Field temperature and moisture removal by the roots effects are neglected.

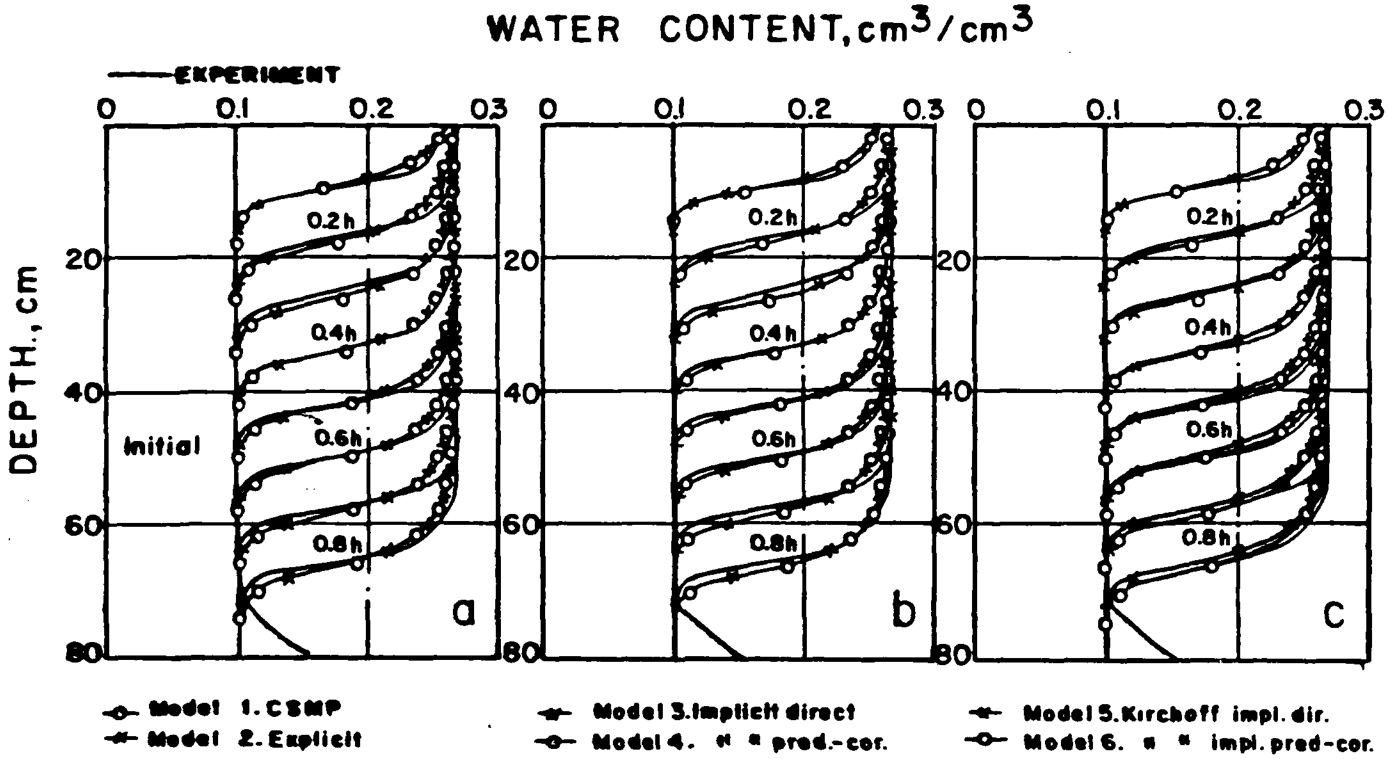

FIG. 6

The results of computations carried out by the different numerical methods for the same problem and for the same values of the parameters as at Fig. 5. Also the experimental data are presented. This picture was taken from [1].

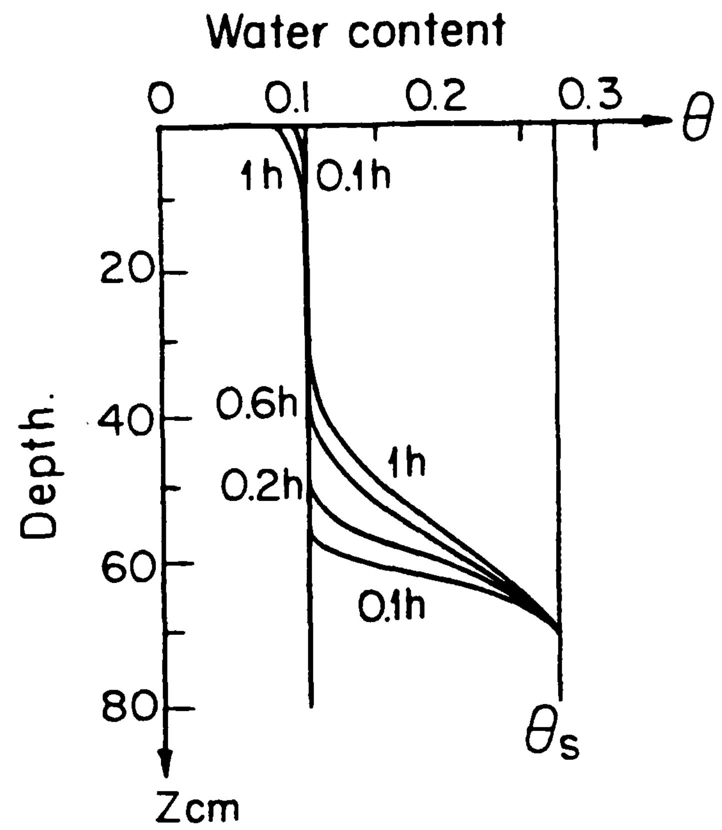

FIG.7

The distribution of ϑ with respect to the depth of the soil for t=0.1h, 0.2h, 0.6h and 1h, computed by our program for the problem W/ρ =0.267 at z=0, ϑ=0.267 at z=70cm, ϑ=0.1 at t=0. (Δz = 1cm, Δt = 5sec) It is assumed

$$K = K_s\left(1 + 0.85\cdot 10^{-6}|\psi|^{4.74}\right)$$

$$\vartheta = (\vartheta_s - \vartheta_d)\cdot\left(1 + 0.62\cdot 10^{-6}|\psi|^{3.96}\right) + \vartheta_d$$

where K =34cm/h, ϑ_s=0.267, ϑ_d=0.075. Field temperature and moisture removal by the roots effects are neglected.

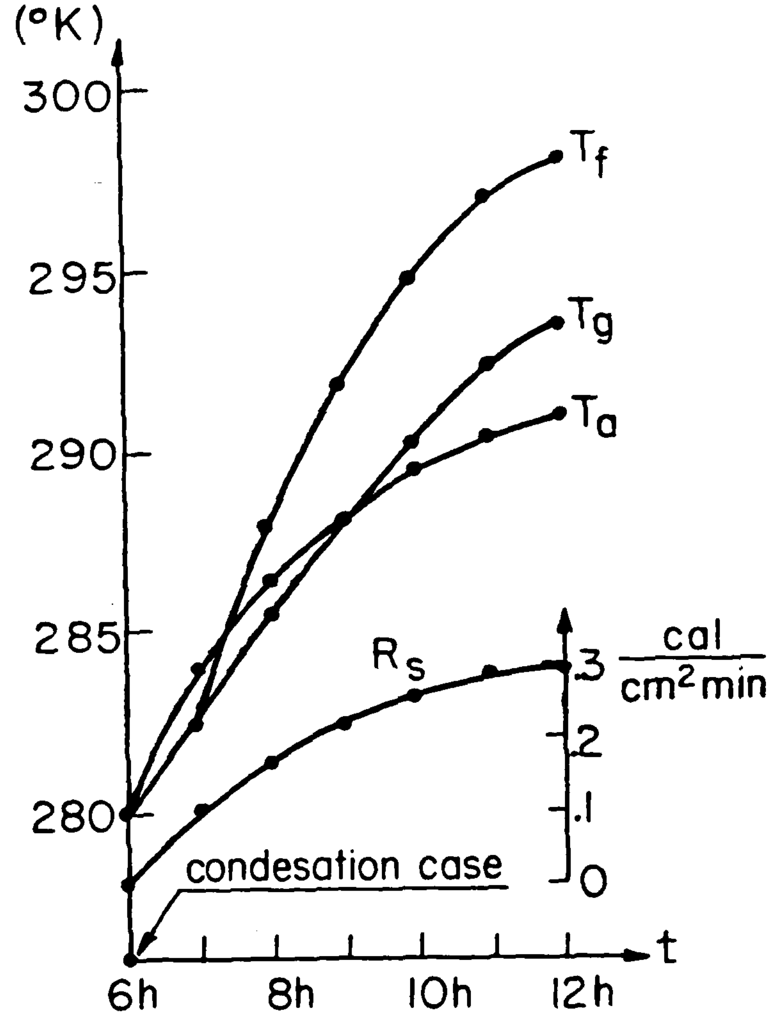

FIG. 8

Calculated diurnal march of T_f and given diurnal marches of T_g , T_a and R_s . It was assumed that q_a =0.007, q_g =0.002+0.005(t^h/6), u =1 m/sec, C_{H_0}=0.057, σ_f=0.75 C_{H_h}=0.096, ϑ_{rmin}=0.2, ϑ_{wilt}=0.1. All this information was taken from [5].

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APPENDIX A.

DETERMINATION OF $(T_f)_j$ AND WHETHER CONDENSATION OR EVAPORATION IS OCCURRING AT THE FOLIAGE AT $t = t_j$.

Setting $r = 1$ in (36), we solve the eq. $F(x, 1) = 0$ numerically by the Newton-Raphson method (see [17]):

$$x^{(m+1)} = x^{(m)} - F(x^{(m)}, 1) \Big/ \frac{\partial F(x^{(m)}, 1)}{\partial x^{(m)}} \qquad m = 0, 1, \ldots \tag{55}$$

The iteration process is stopped when

$$\left| x^{(m+1)} - x^{(m)} \right| < 0.1 \tag{56}$$

The fact that $\partial F(x,1)/\partial x$ and $\partial^2 F(x,1)/\partial x^2$ do not change their signs at $x > 0$ provides the convergence of the iterations for any $x^{(0)} > 0$. After determining x , we compute $\mathcal{B}(x)$ by (37).

Further computation depend on the values of .

If $\mathcal{B}(x) < 0$, then in accordance with (42) condensation on the foliage is occurring and $T_f = x$.

If $\varepsilon_b > \mathcal{B}(x) > 0$, where ε_b is very small (we put $\varepsilon_b = 10^{-5}$), evaporation from the foliage is occurring. However, in this case, the evaporation is sufficiently small to neglect it by putting $E_f = 0$. Then we find ζ from (43) and then determine r from (16) and (17). (Owing to (2), E_{tr} is negligibly small as well as E_f.) In this case T_f is so close to x , that it is possible to take $T_f = x$.

If $\mathcal{B}(x) > \varepsilon_b$, then evaporation from the foliage is occurring. We denote

$\chi = T^{(1)}$. One can show that then $T_f^{(1)} < T_f < T_f^{(2)}$, where $T_f^{(2)}$ is a solution of the eq. $F\left(T_f^{(2)}, \frac{\xi}{1+\xi}\right) = 0$. We turn again to the Newton-Raphson method, finding $T_f^{(2)}$ from the same iterative formula (55) in which one should substitute $\xi/(1+\xi)$ instead of 1, as a second argument of the function F. But now it is better to set $\chi^{(0)} = T_f^{(1)}$.

A criterion for stopping the iterations is the same as (56).

Proceeding to the solution of the eq. (47), we note that the function

$$f(T_f) = \vec{\xi}(T_f) - A \cdot F(T_f, \xi/(1+\xi)) - B \tag{57}$$

is continuous and increasing at the interval $T_f^{(1)} < T_f < T_f^{(2)}$. With the help of (21), (30), (44) and (48) one can show that

$$\left.\begin{aligned} f(T_f) &= 1 - B + \Delta t \cdot a_E \sigma_f \mathcal{E}(T_f^{(1)}) / \vartheta_{d\,max} > 0 \\ f(T_f^{(2)}) &= -B < 0 \end{aligned}\right\} \tag{58}$$

Consequently, since $f(T_f)$ is a monotonic function, the solution of the eq. (47), e.g. $f(T_f^{(2)}) = 0$, exists and is unique. To find this solution, we use the so-called "rule of false position or reguli falsi" [19], which is presented by the following convergent iteration process

$$T_f^{(n+1)} = \begin{cases} \dfrac{T_f^{(n)} f^{(1)} - T_f^{(1)} f^{(n)}}{f^{(1)} - f^{(n)}} & \text{if } f^{(n)} < 0 \\[2ex] \dfrac{T_f^{(2)} f^{(n)} - T_f^{(n)} f^{(2)}}{f^{(n)} - f^{(2)}} & \text{if } f^{(n)} > 0 \end{cases} \tag{59}$$

where we denote $f(T_f^{(n)}) = f^{(n)}$.

The criterion of stopping the iterations is the same as (56).

APPENDIX B

NUMERICAL SCHEME FOR SOLVING THE TEMPERATURE EQUATION (50).

Since C and λ do not depend on T, the eq. (50) is linear. The boundary conditions (27) and (53) are also linear. Therefore, one can use the well known Crank-Nicolson scheme, which is absolutely convergent, to provide an accuracy of the order of $(\Delta t)^2+(\Delta z)^2$ and is amenable for solution via the so-called tridiagonal algorithm [19].

Supplementing the time-grid introduced earlier, we shall introduce a uniform space-grid $z=\Delta z\cdot i$, $i=0,1,\ldots N$ and denote $\varphi(z_i,t_j)=\varphi_{i,j}$, where φ is any variable of the problem.

Presenting the eq. (50) in the form

$$\frac{c}{\lambda}\cdot\frac{\partial T}{\partial t}=\frac{\partial^2 T}{\partial z^2}+\frac{\partial \ln\lambda}{\partial z}\cdot\frac{\partial T}{\partial z} \tag{60}$$

we replace it by the finite-difference equation in accordance with the Crank-Nicolson scheme and then transform it to the following form:

$$A_i\cdot T_{i-1,j}-G_i\cdot T_{i,j}+B_i\cdot T_{i+1,j}=-F_i\quad(i=1,\ldots N-1) \tag{61}$$

where

$$\left.\begin{aligned}
A_i&=1-0.25\ln\left(\lambda_{i+1,j}/\lambda_{i-1,j}\right)\\
B_i&=1+0.25\ln\left(\lambda_{i+1,j}/\lambda_{i-1,j}\right)\\
G_i&=2\left[1+(\Delta z)^2 c_{i,j}/\Delta t\cdot\lambda_{i,j}\right]\\
F_i&=A_i\cdot T_{i-1,j-1}-2\left[1-(\Delta z)^2/\Delta t\cdot\lambda_{i,j}\right]\cdot T_{i,j-1}+B_i\,T_{i+1,j-1}
\end{aligned}\right\} \tag{62}$$

Eq. (61) is solved by tridiagonal algorithm (see [19]):

$$\left.\begin{aligned}&\left.\begin{aligned}\alpha_{i+1} &= B_i/(G_i - \alpha_i A_i)\\ \beta_{i+1} &= (A_i\beta_i + F_i)/(G_i - \alpha_i A_i)\end{aligned}\right\} i = 1,2,..N-1\\ &T_{i,j} = \alpha_{i+1} \cdot T_{i+1,j} + \beta_{i+1} \quad i = N-1,\ldots 1,0\end{aligned}\right\} \tag{63}$$

To employ this algorithm one needs to know α_1 , β_1 and $T_{N,j}$.

The latter value is known from (27) $T_{N,j} = T_d$.

To find α_1 and β_1 , we have to rewrite (53) in the finite-difference form. Remember that the Crank-Nicolson scheme provides accuracy of the order of $(\Delta Z)^2$ with respect to Z . Therefore, we use the following representation:

$$\partial T/\partial Z\Big|_{Z=0} = (3T_0 - 4T_1 + T_2)/2\Delta Z \tag{64}$$

providing the same accuracy.

Substituting (64) into (53), we come to a certain linear equation linking T_0 , T_1 and T_2 . The second linear equation linking these three values is (61) at $i = 1$. Elimination of T_2 from these two equations yields

$$T_{0,j} = \alpha_1 \cdot T_{1,j} + \beta_1 \tag{65}$$

where $\alpha_{1,j}$ and $\beta_{1,j}$ are values of interest:

$$\alpha_1 = (4B_1 - G_1) \Big/ \left[\left(3 + 2\Delta z \cdot \nu/\lambda_{0j}\right)B_1 - A_1\right] ,$$

$$\beta_1 = \left(2\Delta z \cdot \mu \cdot B_1 / \lambda_{0j} + F_1\right) \Big/ \left[\left(3 + 2\Delta z\, \nu/\lambda_{0j}\right)B_1 - A_1\right] . \qquad (66)$$

Note, that in the expressions (54) for ν and μ the surface temperature T_g refers to the previous iteration step. Therefore, the computation $T_{i,j}$ by (63) can be considered as a realization of one iteration step, e.g. as obtaining $T^{(n+1)}$.

Substituting the value $T_{g,j}$ obtained from (63), into (54), we proceed to the next iteration step, etc. As a criterion for stopping the iterations we use the inequality

$$\max\left|T_{i,j}^{(n+1)} - T_{i,j}^{(n)}\right| < 0.1 \; . \qquad (67)$$

APPENDIX C

NUMERICAL SCHEME FOR SOLVING THE SOIL-MOISTURE EQUATION (51).

First, instead of we introduce a new independent variable ("Kirchhoff transformation").

$$v = (\vartheta/\vartheta_s)^{b_T+3} \tag{69}$$

The eq. (51) takes the form:

$$\frac{\partial^2 v}{\partial z^2} = \frac{\vartheta_s}{(-\Psi_s)K_s b_T} \cdot v^{-\frac{b_T+2}{b_T+3}} \cdot \frac{\partial v}{\partial t} + \frac{2b_T+3}{(-\Psi_s)b_T} v^{\frac{b_T}{b_T+3}} \cdot \frac{\partial v}{\partial z} - \frac{b_T+3}{(-\Psi_s)b_T K_s}\left(D_T \cdot \frac{\partial^2 T}{\partial z^2} - S\right). \tag{69}$$

The terms in brackets are known, because T for the moment j has been found and in the expression (6) for S soil moisture ϑ is taken from the $j-1$ moment.

Eq. (69) is more convenient for the numerical solution than the eq. (51) inasmuch as powers in coefficients are not so great, although the parameter b_T can be of order of 10 (see Table 2) from [12].

In addition, the second order derivative term is transformed into its simplest form. As was stated earlier, we use the Douglas and Jones predictor-corrector method [20] for the solution of eq. (69).

First of all, we replace (69) with the system of two finite-difference equations (the first of which is the predictor, and the second is the corrector) following [19] (eqs. (3-50a) (3-50b) in [19]). Then both of these eqs. are reduced to the standard tridiagonal form.

The predictor is given by

$$A_i v_{i-1,j-\frac{1}{2}} - G_i \cdot v_{i,j-\frac{1}{2}} + B_i \cdot v_{i+1,j-\frac{1}{2}} = -F_i \tag{70}$$

where

$$\left.\begin{aligned}
&A_i = 1 \qquad B_i = 1 \qquad G_i = 2(1 + h_{i,j-1}) \\
&F_i = 2h_{i,j-1} \cdot v_{i,j-1} + \frac{1}{2} g_{i,j-1} \cdot (v_{i-1,j-1} - v_{i+1,j-1}) + f_{i,j} \\
&h_{i,j} = \frac{(\Delta z)^2 \vartheta_s}{(-\Psi_s) b_T K_s \Delta t} \cdot v_{i,j}^{-\frac{b_T+2}{b_T+3}} ; \qquad g_{i,j} = \frac{(2b_T+3)\Delta z}{(-\Psi_s) b_T} \cdot v_{i,j}^{\frac{b_T}{b_T+3}} \\
&f_{i,j} = \frac{b_T+3}{(-\Psi_s) b_T K_s} \left[D_T \left(T_{i-1,j-\frac{1}{2}} - 2 \cdot T_{i,j-\frac{1}{2}} + T_{i+1,j-\frac{1}{2}} \right) - (\Delta z)^2 \cdot S \right.
\end{aligned}\right\} \tag{71}$$

and is followed by the corrector:

$$A_i v_{i-1,j} - G_i v_{i,j} + B_i \cdot v_{i+1,j} = -F_i \tag{72}$$

where

$$\left.\begin{aligned}
&A_i = 1 + \frac{1}{2} g_{i,j-\frac{1}{2}} ; \quad B_i = 1 - \frac{1}{2} g_{i,j-\frac{1}{2}} ; \quad G_i = 2\left(1 + h_{i,j-\frac{1}{2}}\right) \\
&F_i = A_i v_{i-1,j-1} - 2\left(1 - h_{i,j-\frac{1}{2}}\right) + B_i v_{i+1,j-1} + 2f_{i,j}
\end{aligned}\right\} \tag{73}$$

To solve (70) and (72) we use the same formula (63) of the tridiagonal

algorithm. It should be noticed that eqs. (70) and (72) are solved successively. First from predictor (70) we find $v_{i,j-\frac{1}{2}}$. Substituting these values into (73), we solve the corrector (72), finding $v_{i,j}$.

The values α_1 and β_1 have to be found from the boundary conditions at the surface (7). To satisfy (7) while computing $v_{i,j}$, we employ the logical scheme, presented in Fig. 4.

It should be noticed that the values $v_{i,j-\frac{1}{2}}$ relate formally to the intermediate time step $t = t_{j-\frac{1}{2}}$, but they can be interpreted as the first approximation to the values of interest $v_{i,j}$. In this regard, the coefficients figuring in the boundary conditions (7) refer to the j-th moment of time in the predictor as well as in the corrector.

The bottom boundary condition (27) gives

$$v_{N,j-\frac{1}{2}} = v_{N,j} = \left\{ \begin{array}{l} v_b < 1 \text{ for dry soil} \\ \text{or} \\ 1 \text{ for saturated soil} \end{array} \right\} \tag{74}$$

One can see that an approximate satisfaction of (7) is achieved at the cost of either the predictor, or the corrector, or both the equations having to be solved twice. At worst, the computations are doubled.

The value $v_{i,j}$ obtained from the corrector is taken as the input value for the next iteration.

The iterations stop when

$$\max \left| v_{i,j}^{(n+1)} - v_{i,j}^{(n)} \right| < 10^{-3} \tag{75}$$

This criterion provides 1% accuracy for θ.

The values α_1 and β_1 corresponding to the conditions No 1 and No 3 follow from the last expression of (63) at $i = 1$, rewritten for $v_{i,j}$. We have $\alpha_1 = 0$, $\beta_1 = 1$. To obtain these parameters for the conditions No 2 and No 4 one must satisfy the moisture balance equation at the surface when the runoff is absent

$$R = P_g - E_g + \rho\left(D_\vartheta \frac{\partial \vartheta}{\partial z} - K + D_T \frac{\partial T}{\partial z}\right)_{z=0} = 0 \qquad (76)$$

rewritten in the finite-difference form for the moments $t_{j-\frac{1}{2}}$ and t_j.

We carried out the following transformation with (76).

1. Introduced v instead of ϑ by (68).
2. Replaced the derivatives by the three-point finite-differences, like (64).

After that the procedure of obtaining α_1 and β_1 is exactly the same as in the case of the soil-temperature equation.

For the condition No 2 we have

$$\alpha_1 = \frac{1 - h_{1,j-1}}{1 + m_{j-1}} \qquad \beta_1 = \frac{M_j + \frac{1}{2}F_1}{1 + m_{j-1}} \qquad (77)$$

where

$$\left.\begin{aligned} m_j &= \frac{(b_T+3)\Delta z}{(-\Psi_s) b_T} \cdot v^{\frac{b_T}{b_T+3}} \\ M_j &= \frac{(b_T+3)\Delta z}{(-\Psi_s) b_T K_s \rho_w}\left(P_g - E_g - X\right) \\ X_j &= \frac{\rho_w D_T}{2\Delta z}\left(3T_{0,j} - 4T_{1,j} + T_{2,j}\right) \end{aligned}\right\} \qquad (78)$$

For the condition No 4 we have

$$\alpha = \left(1 - h_{1,j-\frac{1}{2}} - g_{1,j-\frac{1}{2}}\right) / \left(1 - g_{1,j-\frac{1}{2}} + B_1 \cdot m_{j-\frac{1}{2}}\right) \tag{79}$$

$$\beta = \left(B_1 M_j + \frac{1}{2} F_{1,j}\right) / \left(1 - g_{1,j-\frac{1}{2}} + B_1 \cdot m_{j-\frac{1}{2}}\right) \tag{80}$$

The expressions for the runoff are obtained in this way:

$$R_k = M_k - \frac{(b_r + 3)\Delta z}{(-\psi_s) b_T} - 1.5 + 2V_{1,k} - 0.5V_{2,k} \quad \left(k = j - \tfrac{1}{2}, j\right) \tag{81}$$

The constant dimensional factor in front of the expression for R_k is dropped. One must determine only the sign of this expression.

APPENDIX D.

TABLE OF THE VARIABLES OF THE PROGRAM GREEN

Physical constants and soil-vegetation-atmosphere data

AD	:	air density
AHC	:	specific heat of air at constant pressure
LATC	:	latent heat of vaporization
GRAV	:	gravity acceleration
RW	:	gas constant for water vapour
SIGMA	:	Stefan-Boltzman constant
MSAT	:	moisture value at saturation
MWILT	:	moisture wilting point
FA	:	foliage albedo
FE	:	foliage emissivity
GE	:	ground emissivity
STRES	:	generalized stomatal resistence
FF	:	area averaged foliage shielding factor
WFMAX	:	maximum liquid water retained by foliage per unit horizontal ground area
TRCB	:	dimensionless heat and/or moisture transfer coefficient for a bore soil
UC1	:	it coincides with u_c (26) when $T_h < T_a$
UC2	:	it coincides with u_c (26) when $T_h > T_a$
ANLEV	:	anemometer height
CANLEV	:	canopy height
SRMAX	:	maximum downward shortwave radiation at height "h"

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ANLEV	:	anemometer height
CANLEV	:	canopy height
SRMAX	:	maximum downward shortwave radiation at height "h"

Variables appearing as input values at a given time:

CF	:	cloud fraction
PR	:	precipitation rate
TA	:	air temperature outside the canopy
TH	:	air temperature at canopy height
TF	:	foliage temperature (only a first evaluation to begin the computation)
TG	:	ground temperature
QA	:	air specific humidity
QG	:	ground specific humidity
UA	:	mean wind velocity outside the canopy
SRHD	:	downward shortwave radiation at height "h"
WFN	:	liquid water retained by foliage per unit horizontal ground area
MRMIN	:	minimum moisture in the root zone
MG	:	moisture at the surface

Computation variables:

DTA	:	time interval in the finite difference equation of conservation of the water on foliage
NITMX	:	maximum number of iterations

Variables appearing as input values at a given time:

CF : cloud fraction

PR : precipitation rate

TA : air temperature outside the canopy

TH : air temperature at canopy height

TF : foliage temperature (only a first evaluation to begin the computation)

TG : ground temperature

QA : air specific humidity

QG : ground specific humidity

UA : mean wind velocity outside the canopy

SRHD : downward shortwave radiation at height "h"

WFN : liquid water retained by foliage per unit horizontal ground area

MRMIN : minimum moisture in the root zone

MG : moisture at the surface

Computation variables:

DTA : time interval in the finite difference equation of conservation of the water on foliage

NITMX : maximum number of iterations

Functions

FQSAT (T) : version of the Clausius-Clapeyron equation (19)

FQG (T, MG) : T: temperature; MG: surface moisture; it gives the value of the specific humidity at the surface (25)

FCECK (T): as defined in (37)

Auxiliary variables

R : fraction of potential evaporation rate from foliage

R1 : unitary value of R

R2 : value of R when no water is retained by the foliage

E1 : it coincides with ε_1, (11)

E2 : it coincides with ε_2, (11)

E3 : it coincides with ε_3, (12)

AUXD : computational constant

RL : downward longwave radiation at height "h"

GA : ground albedo

UC : wind speed it coincides with M, (26)

RIB : bulk Richardson number

TRCH : dimensionless heat and/or moisture transfer coefficient at the top of a dense canopy

UAF : mean wind velocity within a canopy

CFO : computational variable

CU : it coincides with C_u, (14)

AUXA : computational variable

AA : it coincides with a_a ,(33)

AF : it coincides with a_f ,(33)

AG : it coincides with a_g ,(33)

AE : it coincides with a_E ,(35)

RE : it coincides with R_E ,(38)

XIAUX : computational variable

XI : it coincides with ξ ,(18)

AUXF : computational variable

A0 : it coincides with A ,(48)

B0 : it coincides with B ,(48)

QAG : it coincides with q_{ag} ,(35)

TAG : it coincides with T_{ag} ,(39)

NITER : number of iterations

CECK : value of $\mathcal{B}(T_f)$, (37)

FAUX : computational variable

F1 : computational variable

F2 : computational variable

FDER : derivative of the function $F(T_f ,r)$,(36) with respect to the temperature

TFP : foliage temperature value referred to the previous iteratixe step.

TF1 : foliage temperature calculated by solving equation (36) with r=1.

TF2 : foliage temperature calculated by solving equation (36) with $r=\frac{\xi}{1+\xi}$ (no water is retained by the foliage)

EF : evaporation rate from foliage; it coincidence with E_f in (40)

FTF1 : value of $f(T_f)$ defined (57) with T=TF1

FTF2 : value of $f(T_f)$ defined (57) with T=TF2

QSAT : saturation specific humidity

QAF : mean specific humidity inside the canopy, (31)

TAF : mean temperature inside the canopy, (32)

RTF : fraction of the potential evapotranspiration rate as a function of T_f, (45)

F3 : value of the function $F(T_f, r)$, (36)

FTF : value of the function $f(T_f)$, (57)

```
      PROGRAM GREEN  (INPUT,OUTPUT)
C
C
C     THIS PROGRAM SOLVES THE ENERGY BALANCE EQUATION FOR A VEGETATION
C     LAYER AS PARAMETERIZED BY DEARDORFF
C
C     UNITS ARE: CM MIN CAL GRAMS
C
C
      REAL LATC,MC,MWILT,MRMIN,MSAT
      DATA GRAV/981./,RW/461.5E6/,LATC/585./,AHC/.24/,AD/1.26E-3/,
     $SIGMA/.826E-10/,FA/.2/,FE/.95/,GE/.95/,STRES/2./,WFMAX/.1/,
     $TRCB/.0057/,R1/1./,MWILT/.1/
C
C     FUNCTIONS*****************************************************
C
      FQSAT(X)=.38E-2*EXP(17.269*(X-273.16)/(X-35.86))
C     FQG(X,Y)=FQSAT(X)*EXP(GRAV*Y/(RW*X))
      FCECK(X)=AHC/LATC*(TAG-X)+1/(LATC*AE)*(RE-E3*X**4)
C
C     **************************************************************
C
C     SET VEGETATION PARAMETERS AND COMPUTATION STEPS
C
      PRINT*,'ASSIGN FF ANLEV CANLEV'
      READ*,FF,ANLEV,CANLEV
      DTA=3
      NITMX=20
C
C     AUXILIARY CONSTANTS
C
      E1=FE*GE*SIGMA/(FE+GE-FE*GE)
      E2=(1-FF)*SIGMA*GE+E1*FF
      E3=E1+SIGMA*FE
      AUXD=17.269*237.3*.38E-2
C
C     INPUT VALUES FOR THE VEGETATION LAYER EQUATION SOLUTION
C
      PRINT*,'ASSIGN CF RR  '
      READ*,CF,RR
      PRINT*,'ASSIGN TA TH TC'
      READ*,TA,TH,TC
      PRINT*,'ASSIGN QA QG'
      READ*,QA,QG
```

```
      PRINT*,'ASSIGN UA'
      READ*,UA
      UC1=600
      UC2=6600
      PRINT*,'ASSIGN SRMAX SRHD'
      READ*,SRMAX,SRHD
      PRINT*,'ASSIGN WFN'
      READ*,WFN
      PRINT*,'ASSIGN MRMIN MG MSAT'
      READ*,MRMIN,MG,MSAT
      TF=TA
      RL=(CF+1.21*(1-CF)*QA**.08)*SIGMA*TA**4
      IF(MG.LE.MSAT) GA=.31-.17*MG/MSAT
      IF(MG.GT.MSAT) GA=.14
      IF(TH.LT.TA) UC=UC1
      IF(TH.GE.TA) UC=UC2
      RIB=GRAV*(ANLEV-FF*CANLEV)*(1-TH/TA)/(UA**2+UC**2)
      IF(RIB.LT.0.) TRCH=AUX1*(1+24.5*(-AUX1*RIB)**.5)
      IF (RIB.GE.0.) TRCH=AUX1/(1+11.5*RIB)
      UAF=(1-FF+.83*FF*TRCH**.5)*UA
      CFO=.01*(1+1800/UAF)
      CU=7.7E-2*(UAF+30)
      TRCG=(1-FF)*TRCB+FF*TRCH
      AUXA=TRCG*UAF+FF*(CU+TRCH*UA)
      AA=TRCH*UA/AUXA
      AF=FF*CU/AUXA
      AG=1-AA-AF
      AE=AD*CU*(1-AF)
      RE=(1-FA)*SRHD+FE*RL+E1*TG**4
      XIAUX=1/(SRMAX/(SRHD+.3*SRMAX)+(MWILT/MRMIN)**2)
      XI=XIAUX/(STRES*CFO*UAF)
      R2=XI/(1+XI)
      AO=DTA*AD*CU*FF*(1-FA+XI)/WFMAX
      BO=WFN+DTA*FF*PR/WFMAX
      QAG=(AA*QA+AG*QG)/(1-AF)
      TAG=(AA*TA+AG*TG)/(1-AF)
      PRINT*,'QAG TAG ',QAG,TAG
      AUXF=XI/(1-AF+XI)
C
C     NEWTON-RAPHSON METHOD FOR THE SOLUTION OF THE CANOPY ENERGY BALANCE
C     EQUATION
C
C     ITERATIVE CALCULATIONS
C
    1 NITER=NITER+1
      IF (NITER.EQ.NITMX) GO TO 7
      CECK=FCECK(TF)
```

```
      FAUX=FOSAT(TF)-OAC
      F1=AUXD*FOSAT(TF)/(TF-35.86)**2
      F2=(AHC+4*E3*TF**3/AE)/LATC
      FDER=F1+F2
      TFP=TF
      TF=TFP-(FAUX-CECK)/(F1+F2)
      IF(ABS(TF-TFP).GT..01) GO TO 1
      PRINT*,'FIRST COMPUTATION TF=',TF
      TF1=TF
C
C     CECK FOR CONDENSATION OR NEGLIGIBLE EVAPORATION CASE
C
      IF(CECK.GE..00001) GO TO 2
C
C     CONDENSATION CASE
C
      PRINT*,'CONDENSATION CASE TF=',TF
      EF=FF*AE*CECK
      WFN=WFN+(FF*PR+EF)*DTA/WFMAX
      IF (WFN.LE.1) GO TO 6
      PRINT*,'WFN EXCEDED'
      GO TO 8
C
C     EVAPORATION CASE
C
    2 R=R2
      NITER=0
      TF1=TF
    3 NITER=NITER+1
      IF (NITER.EQ.NITMX) GO TO 7
      CECK=FCECK(TF)
      FAUX=AUXF*(FOSAT(TF)-OAG)
      F1=AUXF*AUXD*FOSAT(TF)/(TF-35.9)**2
      F2=(AHC+4*E3*TF**3/AE)/LATC
      FDER=F1+F2
      TFP=TF
      TF=TFP-(FAUX-CECK)/FDER
      IF(ABS(TF-TFP).GT..01) GO TO 3
      PRINT*,'EVAPORATION CASE FIRST EVALUATION OF TF: ',TF
      TF2=TF
C
C     METHOD OF RULE OF 'REGULI FALSI' FOR THE SOLUTION OF THE FOLIACE
C     MOISTURE CONSERVATION EQUATION
```

```
C
      IF (BO.LT..00001) GO TO 5
      NITER=0
      FTF1=1-BO+DTA*AE*FF*FCECK(TF1)/WFMAX
      FTF2=-BO
      TF=(TF1+TF2)/2
 4    NITER=NITER+1
      IF (NITER.EQ.NITMX) GO TO 7
      CECK=FCECK(TF)
      QSAT=FQSAT(TF)
      QAF=QAG+AF*CECK
      RTF=(1-AF)*CECK/(QSAT-QAF)
      WFN=((XI+1)*RTF-XI)**1.5
      IF (WFN.LE.1.) GO TO 44
      PRINT*,'WFN EXCEDED'
      GO TO 8
 44   F3=AUXF*(QSAT-QAG)-CECK
      FTF=WFN-AO*F3-BO
      TFP=TF
      IF(FTF.GT.0.) TF=(TF2*FTF-TFP*FTF2)/(FTF-FTF2)
      IF(FTF.LT.0) TF=(TFP*FTF1-TF1*FTF)/(FTF1-FTF)
      IF (ABS(TF-TFP).GT..1) GO TO 4
 5    PRINT*,'EVAPORATION CASE TF=',TF
 6    TAF=AA*TA+AF*TF+AG*TG
      QAF=QAG+AF*FCECK(TF)
      PRINT*,' TAF QAF ',TAF,QAF
      TH=(1-FF)*TG+FF*TAF
      PRINT*,' VALUES MEMORIZED FOR SUCCESSIVE BALANCE EVALUATION'
      PRINT*,'TIME TF TH WFN ',TIME,TF,TH,WFN
C
      PRINT*,'TIME TG TA TF  ',TIME,TG,TA,TF
      GO TO 8
 7    PRINT*,'NITER EXCEDED ',NITER
 8    CONTINUE
      STOP
      END
END OF FILE
```

APPENDIX E.

TABLES OF THE VARIABLES OF THE PROGRAM SOIL

Soil data

TETS	:	Volumetric water content at the saturation.
TETR	:	residual volumetric water content
TETN	:	initial constant volumetric water content
KS	:	saturated soil hydraulic conductivity
B	:	it coincides with β in the first of (12)
AI	:	it coincides with A^{-1} in the first of (12)
BETA	:	it coincides with β in the second of (12)
ALFA	:	it coincides with α^{-1} in the second of (12)

Computation variables

N	:	number of layers
N1	:	number of levels
DZ	:	depth of a layer
DT	:	time step
JMAX	:	2T/DT where T is the duration of the infiltration process
IPRINT	:	index for printing
J	:	computation step: there are two computation steps for each time step

Variables computed at each iteration

H(I)	:	soil water pressure
TET(I)	:	volumetric water content
K(I)	:	soil hydraulic conductivity
C(I)	:	soil water capacity
CK(I)	:	soil water diffusivity
HOLD (I)	:	soil water pressure referred to the pevious time step.
F(I)	:	element of the tridiagonal matrix
G(I)	:	element of the tridiagonal matrix
AC(I)	:	coefficient of the tridiagonal algorhythm
BC(I)	:	coefficient of the tridiagonal algorhythm
TIME	:	time

Anxiliary variables

FACT	:	computational auxiliary variable
AUX1	:	computational auxiliary variable
AUX2	:	computational auxiliary variable
AUX3	:	computational auxiliary variable

Functions

FK(H)	:	soil hydraulic conductivity as a function of the water pressure
FC(H)	:	soil water capacity as a function of the water pressure
FTET(H)	:	soil water characteristic function.

These are expresions (12) in [1]

```
      PROGRAM SOIL  (INPUT,OUTPUT)
C
C     ONE-DIMENSIONAL INFILTRATION MODEL
C
C     HAVERKAMP SOIL FUNCTIONS: CASE: SAND
C
C     UNITS ARE: CM HOUR
C
C
      REAL KS
      DOUBLE PRECISION FK,FC,FTET,K,C,CK,F,G,AC,BC,H,HOLD,AUX1,AUX2,
     $AUX3
      DIMENSION H(81),K(81),C(81),CK(81),G(80),F(80),AC(80),BC(80),
     $HOLD(81),TET(81)
C
C
      DATA TETR/0.075/,TETS/0.287/,ALFI/0.621E-6/,BETA/3.96/,
     $AI/0.851E-6/,Q/13.69/,DZ/1./,TETN/0.1/,KS/34./,
     $B/4.74/,HN/-61.5/,N/70/,H1/-20.73/
C
C     SOIL FUNCTIONS**************************************************
      FK(X)=KS/(1+AI*ABS(X)**B)
      FC(X)=(TETS-TETR)*ALFI*BETA*ABS(X)**(BETA-1)/
     $(1+ALFI*ABS(X)**BETA)**2
      FTET(X)=(TETS-TETR)/(1+ALFI*ABS(X)**BETA)+TETR
C
C    *****************************************************************
C
C
      PRINT*,'ASSIGN JMAX'
      READ*,JMAX
      DT=1./720
      N1=N+1
      FACT=DZ**2/DT
C
C     BOUNDARY CONDITION, IMPOSED FLUX: IT=2;  FIXED VALUE: IT=3
C
      IT=2
C     IT=3
C
C     INITIAL CONDITIONS
C
      DO 10 I=1,N1
      H(I)=HN
      HOLD(I)=HN
      K(I)=FK(HN)
      C(I)=FC(HN)
      TET(I)=FTET(HN)
   10 CK(I)=C(I)/K(I)
C
```

```
C       FIRST LEVEL WHEN A FLUX IS IMPOSED
C
        PRINT*,'ASSIGN HI'
        READ*,HI
        H(1)=HI
        HOLD(1)=HI
        K(1)=FK(HI)
C
C       FIRST LEVEL FIXED
C
C       H(1)=H1
C       HOLD(1)=H1
C
        PRINT*,'INITIAL TETA  PROFILE'
        PRINT 100,(I,TET(I),I=1,N1)
C       PRINT*,'INITIAL PRSSURE HEAD'
C       PRINT 100,(I,H(I),I=1,N1)
C       PRINT*,'INITIAL VALUES K C CK'
C       PRINT 200, (I,K(I),C(I),CK(I),I=1,N1)
C
        J=0
        IPRINT=1
        TIME=0.
C
C       BEGINNING TIME LOOP
C
      1 J= J+1
        IF(J/2*2.NE.J) GO TO 2
        TIME=TIME+DT
C
        DO 20 I=2,N
        AUX1=(K(I+1)-K(I-1))/K(I)
        AUX2=(H(I+1)-H(I-1))/2
        AUX3=HOLD(I+1)-2*HOLD(I)+HOLD(I-1)
     20 F(I)=2*CK(I)*FACT*HOLD(I)+AUX1*(AUX2-DZ)+AUX3
C
        GO TO 3
C
      2 DO 30 I=2,N
        AUX1=(K(I+1)-K(I-1))/K(I)
        AUX2=(H(I+1)-H(I-1))/2
     30 F(I)=2*CK(I)*H(I)*FACT+AUX1*(AUX2-DZ)/2
C
      3 CONTINUE
C
        DO 40 I=2,N
40      G(I)=2*(1+FACT*CK(I))
C
```

```
C       FIXED VALUE AS BOUNDARY CONDITION
C
C       AC(IT)=1/G(2)
C       BC(IT)=(H1+F(2))/G(2)
C
C       IMPOSED FLUX AS BOUNDARY CONDITION
C
        AC(IT)=2-G(2)/2
        BC(IT)=F(2)/2+DZ*(Q/K(1)-1)
C
        DO 50 I=IT+1,N1
        AC(I)=1/(G(I-1)-AC(I-1))
   50   BC(I)=(BC(I-1)+F(I-1))/(G(I-1)-AC(I-1))
C
C       IN THE CASE OF A FIXED VALUE
C
C       DO 60 I=1,N-1
C
C       IN THE CASE OF IMPOSED FLUX
C
        DO 60 I=1,N
        I1=N-I+1
        HOLD(I1)=H(I1)
        H(I1)=AC(I1+1)*H(I1+1)+BC(I1+1)
        K(I1)=FK(H(I1))
        C(I1)=FC(H(I1))
        TET(I1)=FTET(H(I1))
   60   CK(I1)=C(I1)/K(I1)
C
C     6 IF(J/2*2.NE.J) GO TO   66
      6 IF(J.EQ.144*IPRINT) GO TO 65
        GO TO 66
C
    65 PRINT*,'TETA PROFILE   TIME=',TIME
       PRINT 100, (I,TET(I),I=1,N1)
C      PRINT*,'PRESSURE HEAD PROFILE'
C      PRINT 100,(I,H(I),I=1,N1)
       IPRINT=IPRINT+1
C      PRINT*,'K   C   CK'
C      PRINT 200, (I,K(I),C(I),CK(I),I=1,N1)
C
   66  IF(J.EQ.JMAX) GO TO 4
       GO TO 1
    4  CONTINUE
  100  FORMAT(/,5(4X,I3,E11.4))
  200  FORMAT(/,I2,3E11.4)
       STOP
       END
--EOR--
END OF FILE
```

DTIC

END

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